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Measuring Forest EVAPOTRANSPIRATION —Theory and Problems

by C. Anthony Federer

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SOME BASIC PROBLEMS

A SATISFACTORY general method of measuring forest evapotranspiration has yet to be developed. Many procedures have been tried, but only the soil-water budget method and the micrometeorological methods offer any degree of success. This paper is a discussion of these procedures and the problems that arise in applying them. It is designed as a reference for scientists and management specialists working in forest hydrology and meteorology.

In forest hydrology, problems that require knowledge of the amount of water being lost to the atmosphere by evaporation and transpiration can be divided into two major groups: (1) estimating evapotranspiration from an existing fully forested surface, and (2) predicting the changes in evapotranspiration that would follow any given treatment. These problems can generally be subdivided according to the time and area scales involved. Water-yield problems may be of concern on an annual, seasonal, monthly, or even weekly basis. Flood-prediction problems may require knowing the changes in soil-water storage by monthly, weekly, or even shorter intervals. The area of the watershed under study can range from a few acres to thousands of square miles.

At some time, apparently still in the distant future, we can expect that a rational procedure for estimating forest evapotranspiration will be developed — based on soil, vegetation, and topographic parameters and soil and atmospheric variables. But we are still searching for basic understanding of the many processes involved in the translocation of water from the soil to the atmosphere. Concurrently, we are trying to develop techniques that can be used to measure evapotranspiration under some limit-

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ing conditions for application to some of the hydrologic problems discussed above.

The standard device for accurate measurement of evapotranspiration from agricultural crops is a weighing lysimeter. Other techniques are generally tested against a lysimeter. Unfortunately, because of the size of forest vegetation, the weighing lysimeter is unsuitable for forest evapotranspiration studies. In the gaged-watershed approach to forest hydrology, the watershed is considered as a large unweighed lysimeter. If precipitation is measured, and if measured streamflow represents all surface and subsurface water loss from the watershed, then the gaged watershed gives evapotranspiration by difference for any period in which soil water storage at the end is the same as at the beginning. This last restriction often limits the method to annual or perhaps to growing-season values. Unmeasured surface and subsurface losses often invalidate this method completely.

The sapflow method, which measures the rate of water movement in the tree stem, cannot yet be used to determine transpiration rates *(Swanson 1967)*. The quick-weighing method of measuring the weight loss of excised leaves or branches has been used by Rutter *(1966)* to obtain forest transpiration, but this method is laborious and gives only instantaneous values. And the method of enclosing a branch or whole tree in a tent and measuring the water vapor gained by the air is unsatisfactory because it alters the transpiration rate. These three methods, which are discussed by Swanson and Lee *(1966)*, are subject to great sampling errors; and they give values of transpiration only.

Empirical techniques have been used to estimate evapotranspiration from measured evaporation loss of a small water surface (an evaporation pan or atmometer) or from one or two meteorological parameters such as air temperature, saturation deficit, or solar radiation. But the applicability of these or any other such methods proposed in the future cannot be tested unless a suitable method of actually measuring forest evapotranspiration can be found.

The application of either the soil-water budget method or the micrometeorological approaches to large watersheds will remain

impractical because of the spatial sampling involved. Evapotranspiration from large watersheds must continue to be evaluated by semi-empirical techniques. However, continual improvement of these techniques must be based on studies involving the soil-water budget and micrometeorological methods on smaller areas. The soil-water budget method, in which the change of soil water content is measured, has been the normal method for experimental plot studies for a long time.

Micrometeorological methods of determining evapotranspiration for crop, soil, and water surfaces have been developed only in the past 15 years. These involve measuring the energy balance of the surface and the turbulent transfer of heat and vapor in the air above the surface. The development of these methods has greatly increased understanding of the physical factors controlling the evaporation process.

THE SOIL-WATER BUDGET METHOD

The familiar water-balance equation provides one approach to measuring forest evapotranspiration *(Burroughs and Schultz 1965)*. In its most general form for any given soil-vegetation volume, this equation[1] can be written

$$\hat{E} = P \pm F \pm G \pm \Delta W \qquad (1)$$

where $\hat{E}$ is the water output above the surface or evapotranspiration; P is the water input above the surface or precipitation; F is the flow in or out on the surface, or streamflow; G is the flow in or out below the surface, or deep seepage; and ΔW is the change in water content of the soil and vegetation volume, and of snow cover. The four terms on the right side of the equation must be measured in order to evaluate $\hat{E}$; but under some conditions any one, two, or three terms may be negligible.

We do not need to discuss the measurement of precipitation and surface runoff here. But note that P must be measured above

[1]See Appendix for a description of the values in equations (1) to (14).

the canopy if interception is to be included in $\hat{E}$. Also, there may be surface inflow in some situations.

The deep seepage term, G, has always been a bugaboo. Any loss or gain of water through the bottom or sides of the soil volume sampled for the $\triangle W$ term is included. If the whole soil volume is relatively dry, deep seepage may be negligible; but movement can occur when the soil is drier than field capacity *(Hewlett and Hibbert 1963)*. When a true water table is present, knowledge of the spatial variation of the height of the water-table surface and the hydraulic conductivity of the soil may be sufficient for estimating the inflow or outflow. Similarly, for soils drier than saturation, the appropriate soil-water suction gradients and the hydraulic conductivity will give, at least in theory, the inflow or outflow *(Holmes 1960; Rose and Stern 1965)*. Practical problems with instrumentation as well as sampling and theoretical difficulties, especially with hysteresis, cannot be overemphasized. Several methods of indirectly estimating deep seepage have also been devised *(Slatyer and McIlroy 1961; Willardson and Pope 1963)*.

Much has been written about measuring soil water storage. Slatyer and McIlroy *(1961)*, *Taylor* et al. *(1961)*, and Cope and Trickett *(1965)* have given critical reviews of the various methods. For some studies, electrical resistance units may be satisfactory *(Rutter 1964)*. However, present practice favors the neutron-scattering technique. The problem of stones in inserting access tubes is being overcome by use of power drills *(Koshi 1966; Richardson 1966)*. Attempts to achieve satisfactory measurements within the surface 30 cm. *(van Bavel 1961)* have led to commercial surface probes and shields for depth probes *(Pierpoint 1966)*, but the usefulness of surface-probe measurements in evapotranspiration studies has yet to be determined.

Hewlett, et al. *(1964)* showed that instrument and timing errors in the neutron-scattering method are insignificant compared to the spatial sampling error for Carolina forest soils. They also showed that in time, repeated sampling at the same points causes the standard error of the change in mean water content to be only one-third of the standard error of the mean absolute water

content. This advantage is not obtained with gravimetric sampling.

Hewlett, et al. *(1964)* obtained a standard error of about 0.3 percent water content for 30 samples over several acres, each having 30-second timing intervals. They showed that further increase of the number of sampling locations or the sampling time would not significantly reduce the standard error. Converting the 30 samples from percent by volume to inches of water, assuming a 75 cm. soil depth, yields 0.22 cm. as their standard error. This is of the same order of magnitude as the daily evapotranspiration, so accurate daily difference in soil-water content appear to be impossible to detect unless the soil is extremely homogeneous. The soil-water balance thus cannot give meaningful values of evapotranspiration for periods of less than 3 or 4 days.

Many studies have been made of the water balance of forests, often to determine differences in evapotranspiration among cover types over weekly periods *(Zahner 1955; Lull and Axley 1958)*. Sometimes inadequate attention is paid to the deep seepage problem, so results must be carefully evaluated. The soil-water balance approach provides information on rooting depth and soil-water content, but it gives no information about the meteorological factors that govern evapotranspiration.

MICROMETEOROLOGICAL APPROACH

Two basic principles underlie the micrometeorological methods of determining evapotranspiration: the change of phase involved in evaporation requires an energy supply, and the removal of water vapor by the atmosphere is associated with a vapor-concentration gradient. Two assumptions are necessary to develop the present theories: the evaporating surface must be considered two-dimensional and horizontal, with solid (or water) below and air above the interface; and the surface must be infinite in extent. These assumptions are required so that the energy and vapor flow is all vertical; there is no net flow in horizontal directions.

Energy Balance

The energy used in evapotranspiration, L_vE, is the product of the latent heat of vaporization, L_v, and the mass loss of water, E. This energy must be supplied from energy sources. In fact, since the ideal surface is only two-dimensional, conservation of energy requires that as much energy reaches the surface as leaves the surface. The other sources and sinks of energy are the net radiant energy gained or lost, Rn (net radiation); heat gained or lost by air above the surface, H (sensible heat); heat gained or lost by soil and vegetation, S; and heat used in metabolism (mainly photosynthesis minus repiration), M. Melting or freezing of snow also acts as a heat sink or source, which will not be considered here. The resulting surface energy-balance equation, when solved for evapotranspiration, is

$$L_vE = Rn - H - S - M \tag{2}$$

All terms in this equation may have units of energy per unit surface area per unit time or, when integrated over a given time period, just energy per unit area. Net radiation is here positive for energy flow to the surface, while other terms are positive for flow away from the surface. For forests on a normal summer day, all terms are then inherently positive. But in other conditions, any one or more of the terms can become negative: L_vE during dew formation, Rn at night, H at night or during advection (see below), S at night or in the autumn, and M at night due to respiration.

Table 1.—*Diurnal energy balance of a young fir forest*[1]

Flux	Time				Daily Total
	1-2 a.m.	7-8 a.m.	12-1 p.m.	5-6 p.m.	
		Calorie cm^{-2} min^{-1}			*Calorie cm^{-2}*
Rn	−0.06	0.55	1.17	0.38	586
L_vE	− .01	.44	.74	.22	386
H	− .02	.06	.38	.18	197
S	− .02	.00	.03	.01	3
M	− .01	.05	.02	−.03	0

[1]Baumgartner (1956).

Table 1 shows the values of these energy-balance components for a fir forest on a clear summer day *(Baumgartner 1956)*. Other types of forest, and, indeed, other agricultural surfaces, show energy balances similar to this. The S and M terms are generally small for closed forests; and they may be measured or estimated with sufficient accuracy, or else ignored.

The net radiation is actually a combination of two types of radiation — solar and thermal — each of which is transferred both toward and away from the surface (downward and upward). Since most of the energy required for evaporation is supplied by solar radiation, R_s, evaporation is more closely related to R_s than to any other single variable. The fraction of R_s that is reflected by the surface is called the albedo, a. Thermal radiation, also called terrestrial, infrared, or longwave radiation, is emitted both downward by the atmosphere, R_{td}, and—in usually larger amounts —upward from the surface, R_{tu}. The upward component is primarily emission from the surface, which is governed by the surface temperature, T_o, but also includes a reflected part of the downward thermal radiation. Combining these radiation fluxes suitably gives the net radiation

$$Rn = (1 - a)\, R_s + \varepsilon\, R_{td} - \varepsilon\, \sigma\, T_o^4 \qquad (3)$$

where ε is the absorptivity or emissivity of the surface for thermal radiation and σ is the Stefan-Boltzmann constant.

Net radiation can be measured directly with a suitable radiometer; but the other two important terms in the energy balance, L_vE and H, are difficult to measure directly for forests. These two are closely related to the turbulent transfer processes in the atmosphere. Thus it is not possible to find evaporation by the energy balance alone; we must consider the aerodynamic properties of the atmosphere.

Aerodynamic Equations

Four important quantities are transferred vertically by movement of the air above any surface: heat, water vapor, momentum, and carbon dioxide. In each case, the flux in question is proportional to the vertical gradient of, respectively, temperature,

water vapor concentration, wind speed, and CO_2 concentration. The CO_2 flux is important in plant growth studies but does not concern us here.

Sensible heat is heat transferred to or away from the surface because the surface is colder or warmer than the air above. Momentum transfer from the air to the surface is due to friction exerted on the air moving over the surface and is evidenced by lower wind speed closer to the surface. Water vapor generally moves away from the wet surface into the drier air as evaporation; but sometimes it moves in the reverse direction, resulting in condensation as dew or frost.

The equations governing these transfer processes are similar:[2]

$$H = - C_p \rho K_h (dT / dz) \tag{4}$$

$$E = - \rho K_w (dq / dz) \tag{5}$$

$$\tau = \rho K_m (du / dz) \tag{6}$$

The sensible heat flux, H, and the water vapor flux, E, appeared above in the energy-balance equation (2). The momentum flux is τ. The density of the air, ρ, and its specific heat, C_p, are considered constants. The transfer coefficients or diffusivities, K_h, K_w, and K_m for heat, water vapor, and momentum, respectively, are not constant but vary with windspeed, surface roughness, height, and sensible heat flux, H. The terms in parentheses represent the vertical gradients or change with height z, of air temperature, T; specific humidity, q; and wind speed, u.

The three transfer coefficients are closely related, since all are dependent on the turbulent properties of the atmosphere. Many experiments have shown that

$$K_m = k u_* z / \phi_m \tag{7}$$

where k is the von Karman constant, ϕ_m is a stability correction,

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[2]The gradients here and later in this report can be written as total derivatives (dT/dz) rather than as partial derivatives ($\delta T/\delta z$) because horizontal homogeneity and steady state conditions are assumed. When this assumption does not hold, advection occurs (see page 00). And to be strictly correct, the adiabatic lapse rate, 0.01°C m^{-1}, should be subtracted from dT/dz.

and u_* is a form of the momentum flux called the friction velocity, which is defined as

$$u_* = (\tau / \rho)^{1/2} \qquad (8)$$

When the sensible heat flux, H, is zero (neutral or adiabatic condition), the stability parameter, ϕ, is unity. This case is the regime of fully forced convection in which turbulence is produced only by friction due to wind movement over the rough surface. When H is positive (unstable or lapse condition), free convection due to rising warm air and sinking cold air increases the turbulence, which effectively decreases ϕ to some value less than unity. When H is negative (stable or inversion condition), the turbulence is damped and ϕ increases to a value greater than unity.

By an analogy called the similarity hypothesis, the other two diffusivities are similarly defined as

$$K_h = k\, u_*\, z / \phi_h$$

$$K_w = k\, u_*\, z / \phi_w$$

Many scientists assume $\phi = \phi_m = \phi_h = \phi_w$, and thus $K = K_m = K_h = K_w$ as a satisfactory first approximation. This is certainly true in neutral conditions when all ϕ values are unity. But the question of equality of the transfer coefficients in diabatic conditions is an unsolved problem in micrometeorology. Several investigations indicate that K_h may be twice as large as K_m in extremely unstable conditions and only half as large in extremely stable conditions *(Drimmel 1968; Record and Cramer 1966).* Researchers had thought that K_w was approximately equal to K_m *(Deacon and Swinbank 1958),* but recent investigations by Dyer *(1967)* indicate that $K_w = K_h$. Until differences are quantitatively established, equality will have to be assumed for practical work.

A second unsolved micrometeorological problem is the form of ϕ under varying stability conditions. Usually ϕ is given as a function of one of four other stability parameters: the Richardson number in either its flux or gradient forms (R_f and Ri), and

the Monin-Obukhov dimensionless height in either its flux or gradient forms (z / L and z / L'). These are defined here for reference:

$$R_f = \frac{-g H}{C_p \rho u_*^2 T (du / dz)} = \frac{K_h}{K_m} \quad Ri = \frac{K_h}{K_m} \frac{g (dT / dz)}{T (du / dz)^2}$$

$$\frac{z}{L} = \frac{-z g k H}{C_p \rho u_*^3 T} = \frac{K_m}{K_h} \frac{z}{L'} = \frac{K_m}{K_h} \frac{z g k (dT / dz)}{u_* T (du / dz)}$$

where g is the acceleration of gravity and T is the air temperature. The terms are further interrelated since

$$R_f \phi_m = z / L$$

All the parameters R_f, Ri, z / L, and z / L' increase in absolute value nearly linearly with height. The Richardson number at any height z can be found by:

$$Ri = \left[\frac{g (T_2 - T_1)}{T (u_2 - u_1)^2} \ln \frac{z_2}{z_1}\right] z$$

with T the mean of T_1 and T_2.

When the transfer coefficient is to be determined from wind and temperature measurements, the form of ϕ must be known. Although theory requires functions of ϕ involving R_f or L, these cannot be used in practice unless H is independently evaluated through use of such techniques as the eddy correlation method (see below). Therefore, practical forms of ϕ are functions of Ri and L' and thus involve implicit assumption of $K_m = K_h$. Monin and Obukhov *(1954)* introduced a linear relation between ϕ and z / L'. However, this relation contained an empirical coefficient that has been found to vary with stability *(Pandolfo 1963; O'Brien 1965)*, so a number of more complex forms have been proposed *(Panofsky 1963; Swinbank 1964; Webb 1965)*. Brutsaert *(1965)* has given an excellent summary of these relations and their use in evaporation measurements. Bernstein *(1966)* and others who have used existing data to compare these forms find that the experimental data is not good enough to show that one

form is better than another. Thus, in practice, nearly any of the forms may be chosen. In cases of extreme stability and instability, no proposed forms will fit the data adequately.

The friction velocity, u_*, can be found from the integrated form of the combined equations 6, 7, and 8:

$$u_* = \frac{k\,(u_2 - u_1)}{\ln\,[\,(z_2 - D)\,/\,(z_1 - D)\,] - \psi} \qquad (9)$$

$$\text{where } \psi = \int_{z_1 - D}^{z_2 - D} \frac{1 - \phi}{z}\, dz$$

and u_1 and u_2 are wind speeds measured at two heights above the canopy, z_1 and z_2. ψ is the integrated form of the stability correction, which depends on z_1 and z_2. The zero-plane displacement, D, is a correction to the measured heights of the sensors above the ground, z_1 and z_2. D is about equal to the general height of the vegetation and allows for the fact that the top of the canopy, and not the ground level, influences the wind above the trees.

If we extrapolate this wind profile toward the surface, u_1 approaches zero as z_1 — D approaches a constant, z_o. The parameter z_o, is called the roughness length and is characteristic of the physical roughness or turbulence-inducing properties of the surface. If z_o and D are known for forests, u_* can be obtained by measuring u at only one height. D and z_o must be found from measuring wind speed at several heights, and very accurate measurements are required *(Robinson 1961; Covey 1962; and Panofsky 1963)*. For forests z_o, is of the order of 1 m., while the zero plane displacement is somewhat less than the height of the trees *(Baumgartner 1956; Rauner 1960)*.

Mass-Transfer or Profile Method

Substituting (9) into (7) and then into (5) and integrating gives the mass-transfer equation for E as

$$E = -\,\rho\, k^2 \frac{(u_2 - u_1)\,(q_2 - q_1)}{\left[\ln\left(\dfrac{z_2 - D}{z_1 - D}\right) - \Psi\right]^2} \qquad (10)$$

This equation can be used directly for finding evapotranspiration. It requires wind and vapor concentration measurements at two heights above the surface and temperature measurements for the stability correction. The measurement accuracies required are high, since the differences are small. Baumgartner *(1956)* found that this method worked well over a forest at night, but it was unsatisfactory during the day because of insufficient measurement accuracy. Deacon and Swinbank *(1958)* have proposed a modification that eliminates the temperature measurements for the stability correction and assumes a constant but undetermined roughness length.

Bulk aerodynamic methods use surface values, $u_1 = 0$, $(z_1 - D) = z_o$, and $q_1 = q_o$, in the form

$$E = -\rho k^2 \frac{u_2}{\left[\ln\left(\frac{z_2 - D}{z_o}\right) - \psi\right]^2}(q_2 - q_o) \qquad \textbf{(11)}$$
$$= f(u_2)\,(q_2 - q_o)$$

This method involves only wind speed at one height but requires constancy of surface roughness. The exacting measurement of z_o must be made at the outset, or else the wind function $f(u_2)$ can be obtained empirically. The stability correction, if not ignored, requires measurement of temperature at two heights. The surface humidity, q_o, cannot be measured directly except over water, where it is equal to the saturation specific humidity at temperature T_o. We should note here that

$$q = \frac{0.622}{p}\,e$$

where p is atmospheric pressure and e is vapor pressure in the same units as p.

Eddy Correlation Methods

Vertical transport of heat and water vapor in the atmosphere is turbulent in nature. The air flow is made up of a series of eddies, which are carried past a fixed point by the horizontal

wind. The vertical component of the wind speed, w', at a point will vary continuously as the eddies pass, alternately being directed upward and downward. When evaporation is occurring, the air moving upward in the eddies will be wetter than the air moving downward. If we define q' as the instantaneous departure of q from its mean value at the point, then the product $w'q'$ is the deviation of the instantaneous water-vapor flow from its mean value. The evaporation is found by averaging over time as shown by the overbar

$$E = \rho \overline{w' q'}$$

An instrument called the Evapotron has been developed to find E by this method *(Dyer and Maher 1965)*. It shows great promise as a research tool but has not yet been tried over forests.

The sensible heat flux is similarly dependent on turbulent flow, but here temperature takes the place of humidity.

$$H = C_p \rho \overline{w' T'}$$

The Fluxatron *(Dyer et al. 1957)* measures H in this way but also has not yet been applied to forests.

Difference Method

The energy balance equation (2) alone can be used to determine E only if the sensible heat flux is found independently; then H is subtracted from measured Rn to give E. I have called this the difference method, as it apparently has no other commonly used name.

The required sensible heat flux can be found by the eddy correlation method, or it can be found by using equations analogous to (10) and (11) for E. Substituting (9) into (7) and then into (4), and integrating, gives the profile equation for H as

$$H = - C_p \rho k^2 \frac{(u_2 - u_1)\ (T_2 - T_1)}{\left[\ln \left(\frac{z_2 - D}{z_1 - D} \right) - \psi \right]^2} \qquad (12)$$

The difficult measurement of $(q_2 - q_1)$ required for (10) is substituted for by the easier measurements of $(T_2 - T_1)$ and Rn. Accuracy is high over surfaces where evaporation is large and H is small.

The bulk aerodynamic equation for sensible heat flux is similar to (11):

$$H = - C_p \rho k^2 \frac{u_2}{\left[\ln \left(\frac{z_2 - D}{z_o} \right) - \psi \right]^2} (T_2 - T_o) \qquad (13)$$

Combining (13) with (2) is the bulk-difference method for evapotranspiration. A modification of this method has been used over a hardwood forest in Russia *(Rauner 1960; Dzerdzeevskii 1963)*. However, no independent checks of the results were made. The surface temperature, T_o, may be measured directly by infrared radiation measurement *(Fuchs and Tanner 1966)*. Comments applied to equation (11) also apply here.

Tanner *(1960b)* has proposed a modification of (12) similar to Deacon and Swinbank's *(1958)* modification of (10).

Bowen Ratio Method

The Bowen ratio is defined as H / L_vE. This ratio can be obtained from (10) and (12) and substituted into (2) to give, after rearrangement,

$$L_vE = \frac{Rn - S - M}{1 + \left[\frac{C_p (T_2 - T_1)}{L_v (q_2 - q_1)} \right]}$$

Use of the Bowen ratio (the term in brackets) eliminates any need for wind measurements, known heights, stability corrections, roughness length, or zero-plane displacement; but the difficult vapor concentration difference is required *(Suomi and Tanner 1958; Fritschen 1966)*. The ratio K_h / K_w should also appear in the bracketed term if it is not assumed equal to unity. The Bowen ratio method has also been called the energy balance method, but this name is misleading. Baumgartner, *(1956)* in his classic work

on a fir forest in Austria, used this method, but no independent check on the evaporation data was obtained.

Combination Methods

Potential evapotranspiration is defined as the evaporation from a surface for which q_o is the saturation specific humidity at T_o. With this assumption, suitable combination of equations (2), (11), and (13) leads to

$$E_p = \frac{(\triangle / C_p)\,(Rn - S - M) + \rho\, h\, (q_{T2} - q_2)}{1 + (\triangle\, L_v / C_p)} \qquad (14)$$

$$h = \frac{k^2\, u_2}{\left[\ln\left(\frac{z_2 - D}{z_o}\right) - \psi\right]^2}$$

where q_{T_2} is the saturation specific humidity at T_2 and $\triangle$ is the slope of the saturation specific humidity-temperature curve *(van Bavel 1966; Tanner and Fuchs 1968)*. With an empirical wind function for h, this is the well-known Penman *(1948)* method.

The "generalized combination method" for actual evaporation developed by Tanner and Fuchs *(1968)* is simply the bulk difference method described above as the combination of (13) and (2).

A commonly used method of estimating E is to determine E_p by (14) and then to use an empirical relationship of E / E_p to available soil water *(Denmead and Shaw 1962; Crawford and Linsley 1963)*. As soon as this relationship is put on a theoretical basis *(Visser 1963; Cowan 1965; Woo et al. 1966)*, the combination method for potential evapotranspiration and soil-water measurement may give an accurate measurement of evapotranspiration.

The combination method and soil-water approach are also the avenues to predicting effects of forest treatments on evapotranspiration, since they include all relevant variables. Micrometeorological studies of forests have already led to analysis of differences in evapotranspiration from different cover types *(Baumgartner 1967; Rauner 1965)*. All the relevant surface variables — albedo, surface temperature, roughness, stomatal resistance, root and soil water properties — as well as all necessary meteorological vari-

ables — will some day be included in a complete quantitative description of the whole soil-plant-atmosphere system.

PRACTICAL DIFFICULTIES

Instrumentation

The measurements needed for any of the micrometeorological methods unfortunately require sensitive instruments that are not as yet standardized. Net radiation can be measured with a variety of different instruments, some of which are commercially available *(Gates 1965; Federer 1967)*. Differences in wind speed, temperature, and vapor concentration between two heights require extremely accurate sensors because the differences are small. The required order of accuracy needed for the differences are 1 cm. per sec. for wind, and 0.02° C. for wet-bulb and dry-bulb temperature. Greater height differences allow lower absolute accuracy in the measurements. Anemometers meeting these requirements can be purchased. Temperature and water vapor sensors of sufficient accuracy are usually built by the investigator, but no commonly used design is available. An outstanding review of such instrumentation is given by Tanner *(1963)*.

If methods utilizing surface values of temperature and vapor concentration are used, the required accuracies are of a magnitude less than those given above. However, these methods require either an empirical wind function or determination of roughness length and zero-plane displacement from accurate wind profiles, and the assumption that these remain constant.

All the methods discussed here require time integration of the meteorological variables. Integration can be done either electronically at the site before recording, or by digital computer on a series of recorded, instantaneous values. Integration times may vary for different studies from 15 minutes up to half a day. Development of satisfactory recording systems is a significant electronics problem, especially for forest measurements where remote operating stations that may be unattended for periods of a week or more may be required. Calibration drift in either the sensors or the electronics is a serious problem for the systems that must

be employed. Frequent interchange of sensors between heights is one way to reduce this problem *(Tanner 1960a)*. Compromises between the drift problem and unattended operation may be needed.

In short, a great deal of work is needed on instrumentation problems before micrometeorological methods can be easily applied to forests.

Advection

The current theory of the micrometeorological evaporation methods assumes a horizontally uniform, two-dimensional surface. If this assumption is not met, there will be a net horizontal transfer of energy into or out of the volume being measured. This horizontal transfer is called advection, and it is not accounted for by the theory. The three-dimensional form of forests and their horizontal variation imply that advection may be a serious problem in applying micrometeorological methods.

Absorption of the horizontal component of solar radiation is exemplified by isolated trees or stands that have a sunny side and a shadow. Hilly topography also produces inhomogeneities in horizontal solar and infrared transfers. Geiger *(1965, pp. 369-381)* and Lee *(1964)* have analyzed these situations. Lateral interception of water droplets as fog drip is an extreme example of horizontal advection of water vapor, but it can also occur in invisible forms.

Advection of sensible heat and water vapor occurs in three forms (fig. 1). Net movement into or out of a stand, or of individual trees below the tree tops, is called the clothesline effect *(Tanner 1957)*. There is no theory for this effect, and its magnitude has never been measured. Local clothesline advection produced by individual tree crowns and canopy openings causes local variation in energy balance and in mass transfer. Spatial integration then becomes necessary and can be achieved either by using several sensors at different locations or by raising the sensors higher above the canopy.

When sensors are elevated to obtain better spatial sampling and to obtain maximum possible vertical differences, a second

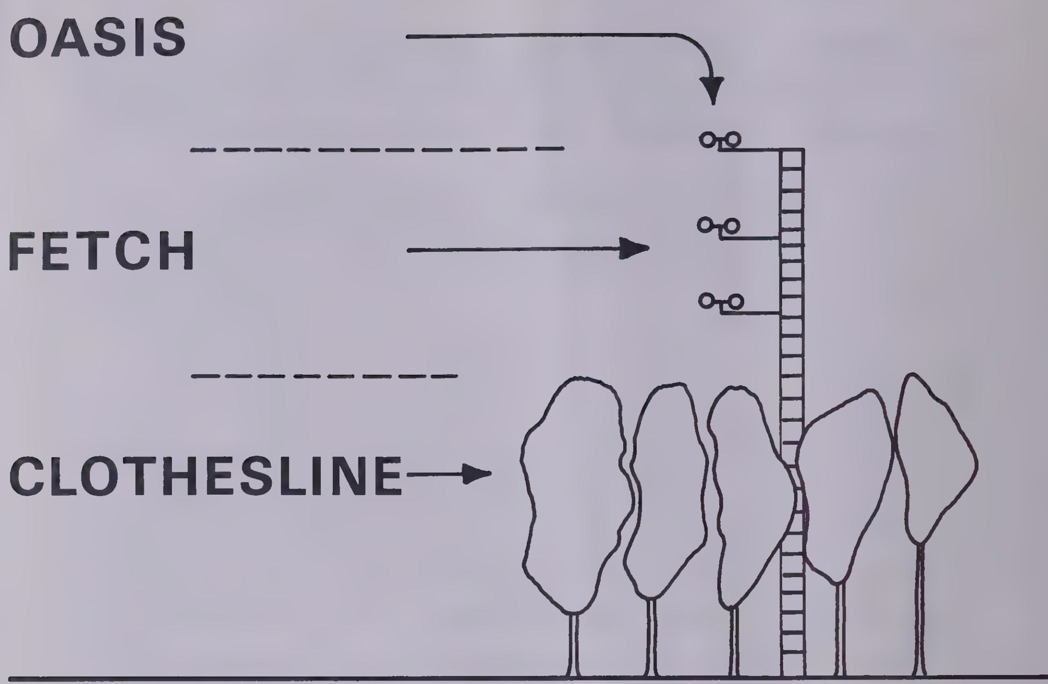

Figure 1.—The three kinds of advection, or net lateral energy transfer.

form of advection becomes important. Energy can be transferred laterally in the space between the top of the canopy and the top of the highest sensing instrument. With increasing height, the wind, temperature, and vapor concentration are affected more and more by surfaces farther upwind and less by the surface immediately below the sensor. The measured vertical gradients are then not representative of the underlying surface. The fetch requirement of a site is the minimum upwind distance at which a significantly different surface may exist without affecting conditions at the sensor height. This type of lateral transfer may thus be called fetch advection.

Several studies have been made of fetch requirements in the case of a dry surface upwind of a wet surface *(de Vries 1959; Philip 1959; Dyer 1963; Rider et al. 1963)*. Fetch advection can also be produced by a change in surface roughness that affects the downwind turbulent transfer coefficients *(Elliott 1958; Brooks 1961; Panofsky and Townsend 1964)*. A rough rule of thumb requires a fetch 50 to 100 times the highest sensor height above

the zero plane. However, no measurements are available to test this rule over forests. The forest meteorologist is seriously squeezed between the minimum height allowed by clothesline advection and the maximum height allowed by fetch advection.

Energy can also be transferred to an area by horizontal transfer above the sensor height, then by vertical transfer to the surface. This third type of advection is called the oasis effect because the condition is most pronounced over desert oases. Oasis advection is satisfactorily measured by micrometeorological methods since the transfer is vertical at sensor height.

Clothesline advection is net horizontal transfer below the tops of the trees. Fetch advection is net horizontal transfer between the tree tops and the sensing instrument height. Oasis advection is net horizontal transfer above sensor height, or above treetop height if there are no sensors. The theoretical equations discussed in this paper are correct only for regions of height in which none of these three advection forms is appreciable; that is, only for regions in which all net transfer is vertical.

A form of vertical heat transfer within the volume being considered must also be mentioned. Studies by Denmead *(1964)* and Knoerr and Gay *(1965)* indicate that radiant energy absorbed at the upper surface of the forest canopy is in part converted to sensible heat, which is transferred downward into the canopy and used in increasing evaporation. The effective source level, and thus the zero-plane displacements for sensible heat, latent heat, and momentum, may be different. The effect of this phenomenon on the various micrometeorological evaporation methods has not yet been studied.

CONCLUSIONS

Forest hydrology research and forest watershed management have been hindered by lack of a satisfactory method for measuring forest evapotranspiration. Soil-water sampling used in the water balance is limited to periods greater than 3 days and is plagued by the problem of unmeasured deep seepage. Micrometerological methods, based on the energy balance and vapor transport, offer an approach to measuring evapotranspiration for

periods as short as 15 minutes. However, only two successful studies applying these methods to forests have been reported *(Baumgartner 1956; Rauner 1960)*.

Several different micrometeorological methods can be used to obtain nearly independent results. Future experiments should evaluate the usefulness and accuracy of these methods. The magnitude of advection error due to surface inhomogeneities, and consequent spatial sampling requirements, need study. Theoretical problems involving equality of the eddy diffusivities and the diabatic correction may not be serious for the accuracy required here. But problems with instrumentation require further research.

Combining micrometeorological theory with results from soil physics and plant physiology should lead to methods of predicting the effects of changes in the soil-plant-atmosphere system on evaporation.

■

APPENDIX

Notation

The units given below represent a consistent set, but they are by no means the only units that are commonly used in the soil-water budget method and the micrometeorological methods. Energy-flux values are more commonly given with time units in minutes rather than seconds, as shown here.

a	albedo	dimensionless
C_p	specific heat of air	($\approx$ 0.24) cal g^{-1} $^0K^{-1}$
D	zero plane displacement	cm
E	evapotranspiration (rate)	g cm^{-2} sec^{-1}
$\hat{E}$	evapotranspiration (quantity) [note $\hat{E} = \frac{1}{\rho_w}\int E\,dt$]	cm
E_p	potential evapotranspiration	g cm^{-2} sec^{-1}
e	vapor pressure	mb
F	streamflow	cm
G	deep seepage	cm
g	acceleration of gravity	($\approx$ 980) cm sec^{-2}
H	sensible heat flux	cal cm^{-2} sec^{-1}
h	wind function	cm sec^{-1}
K	diffusivity	cm^2 sec^{-1}
k	von Karman constant	($\approx$ 0.42) dimensionless
L	Monin-Obukhov flux length	cm
L′	Monin-Obukhov gradient length	cm
L_v	heat of vaporization	($\approx$ 580) cal g^{-1}
M	heat for metabolism	cal cm^{-2} sec^{-1}
P	precipitation	cm
p	atmospheric pressure	mb
q	specific humidity (g water / g air)	dimensionless
q′	instantaneous deviation from mean specific humidity	dimensionless
q_{T2}	saturation specific humidity at temperature T_2	dimensionless
R_f	flux Richardson number	dimensionless
Ri	Richardson number	dimensionless
R_{td}	downward thermal radiation	cal cm^{-2} sec^{-1}
R_{tu}	upward thermal radiation	cal cm^{-2} sec^{-1}
Rn	net radiation	cal cm^{-2} sec^{-1}
R_s	downward solar radiation	cal cm^{-2} sec^{-1}
S	change in soil or vegetation heat storage	cal cm^{-2} sec^{-1}
T	temperature	^oK
T′	instantaneous deviation from mean air temperature	^oK
u	mean wind speed	cm sec^{-1}
u_*	friction velocity	cm sec^{-1}

w'	instantaneous vertical component of wind	cm sec^{-1}
z	height	cm
z_o	roughness length	cm
Δ	slope of saturation specific humidity-temperature curve	$^oK^{-1}$
ΔW	change in soil water content	cm
ε	thermal emissivity	dimensionless
ρ	air density	(≈ 0.0012) g cm^{-3}
ρ_w	water density	(≈ 1.00) g cm^{-3}
σ	Stefan Boltzmann constant	($\approx 5.7 \times 10^{-5}$) g sec^{-3} $^oK^{-1}$
τ	momentum flux	g cm^{-1} sec^{-2}
ϕ	diabatic correction	dimensionless
ψ	integrated diabatic correction	dimensionless

Subscript $_o$ refers to the surface; $_1$ and $_2$ refer to heights above the surface.

Subscripts $_m$, $_h$, and $_w$ on ϕ and K refer to momentum, heat, and water vapor, respectively.

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